

209925  
p. 37

# *Towards Coupled Physical-Biogeochemical Models of the Ocean Carbon Cycle*

Stephen R. Rintoul

## **Introduction**

The purpose of this review is to discuss the critical gaps in our knowledge of ocean dynamics and biogeochemical cycles. I will start with the assumption that our ultimate goal is the design of a model of the earth system that can predict the response to changes in the external forces driving climate. The prediction aspect is important; simplifying tricks that can be used to describe the present biogeochemical state of the ocean may not be sufficient for predicting future changes. It will also become clear that the design of such a model depends on the time scale it is intended to simulate.

The ocean circulation plays two direct roles in the climate system. First, the poleward flow of warm water and equatorward flow of colder water at depth in the thermohaline circulation results in a heat flux from equator to pole that is roughly equal to that of the atmosphere. Second, the ocean has a large heat capacity relative to that of the atmosphere, so that the timing of any change in global temperatures due to external forcing is largely determined by the thermal inertia of the ocean. Consequently, changes in ocean circulation can lead directly to changes in climate.

The ocean's ability to store and transport heat is perhaps its most important feature with regard to the physical climate system. However, the aim of this review is to focus on those aspects of the climate system that involve the interaction between ocean dynamics and marine biogeochemistry. In particular, I will try to identify those aspects of marine biogeochemistry that need to be included in the

ocean component of a coupled model of the earth system if we are to obtain useful predictions of future climatic change.

To focus the discussion even further, I will limit myself to consideration of the ocean carbon cycle. Carbon participates in active organic and inorganic chemical systems in the ocean, and most other biogeochemical cycles are linked to that of carbon by the marine biota. In addition, much effort has been devoted to modeling the carbon cycle due to concern over the climatic impact of the anthropogenic input of carbon dioxide.

The discussion here, however, will not be limited to the oceanic uptake of anthropogenic  $\text{CO}_2$ . Rather, I will assume that we have a broader interest in understanding how and why the carbon cycle functions as it does at present, and in learning its capacity for change over a range of time scales, from seasons to glacial cycles. The goal is not a comprehensive review, but a personal view of issues that need to be considered to move toward useful, predictive coupled models of the earth system.

Implicit in most carbon cycle modeling to date is an assumption that the system was in steady state prior to the anthropogenic invasion and has remained so in postindustrial times. In particular, the input of  $\text{CO}_2$  due to fossil fuel burning and deforestation is often treated as a small perturbation to the equilibrium carbon system. This is despite the fact that the increase in atmospheric  $\text{CO}_2$  since industrialization is already half that since the last glaciation, when evidence suggests that the ocean circulation and carbon system may have been very different than they are today. To be confident of our model predictions, we need to move beyond the assumption that the carbon system will remain in steady state. This requires a model of how the ocean circulation and carbon cycle interact.

In the following sections, I will review some of the coupled biological-physical models that have been used to date. The models generally represent two extremes of scale. Highly simplified box models have been used to study global phenomena on long time scales, such as the changes in climate between the glacial and interglacial. At the other extreme, more detailed models that explicitly resolve the nonlinear interactions between the biological and physical systems have focused on limited regions and short time scales. By reviewing the results of both types of models, I hope to shed some light on the issue of the most appropriate carbon cycle model for climate studies. In particular, what biological-physical interactions must be included to reproduce the essential processes on the time scales of interest? Modeling is by nature an art of compromise, and the appropriate balance between faithful reproduction of system dynamics and the simplifications that allow increased understanding will depend on the problem considered.

## Oceanic Uptake of Carbon

The uptake of atmospheric  $\text{CO}_2$  by the ocean is limited by several factors: the exchange of  $\text{CO}_2$  across the air-sea interface, the buffer capacity of the oceanic carbonate system, and the transfer between the surface layers of the ocean and the deep sea. In this section, these fundamental aspects of the oceanic uptake of carbon are briefly reviewed.

### Gas Exchange

The exchange of  $\text{CO}_2$  between the atmosphere and the ocean can be expressed as the product of the difference in the partial pressure of  $\text{CO}_2$  between the air and the surface ocean ( $\delta p\text{CO}_2$ ) and a gas transfer coefficient. The gas transfer coefficient increases with increasing wind speed, but the exact form of the wind speed dependence is not well known; present formulations give an uncertainty of about a factor of two in the transfer coefficient (Liss and Merlivat, 1986; Tans et al., 1990). Even if we knew the wind speed-dependence of the transfer coefficient exactly, we would still have problems, because we do not know the wind distribution very well over large parts of the ocean, and we have even fewer observations of oceanic  $p\text{CO}_2$ . Consequently we cannot even "predict" the present exchange of  $\text{CO}_2$  between the ocean and the atmosphere very accurately. Over large areas we are uncertain whether the ocean is a net source or sink.

If we knew the winds and the dependence of the transfer coefficient on wind speed, the problem would be reduced to one of predicting  $p\text{CO}_2$  in the surface ocean. The partial pressure of  $\text{CO}_2$  in the ocean is primarily a function of temperature and the concentrations of alkalinity (ALK) and total dissolved inorganic carbon (DIC).

Due in part to the temperature dependence of the solubility of  $\text{CO}_2$  in sea water, the dominant factor in determining  $\delta p\text{CO}_2$  is ocean upwelling and downwelling (Keeling, 1968). Upwelled water warms at the sea surface and releases  $\text{CO}_2$ ; water cooled at high latitudes takes up  $\text{CO}_2$ . Maps of oceanic  $p\text{CO}_2$  from the scarce data presently available agree with this broad-brush picture of the ocean as a sink for atmospheric  $\text{CO}_2$  at high latitudes and a source at low latitudes (Figure 1).

The temperature dependence of  $p\text{CO}_2$  implies that the physical factors that determine the temperature of the mixed layer also affect  $p\text{CO}_2$ . Therefore, to be able to predict oceanic  $p\text{CO}_2$  we need to model the physical mechanisms that affect temperature (lateral and vertical advection and mixing, and heating and evaporation at the sea surface) and the biochemical mechanisms that determine the concentrations of alkalinity and DIC.

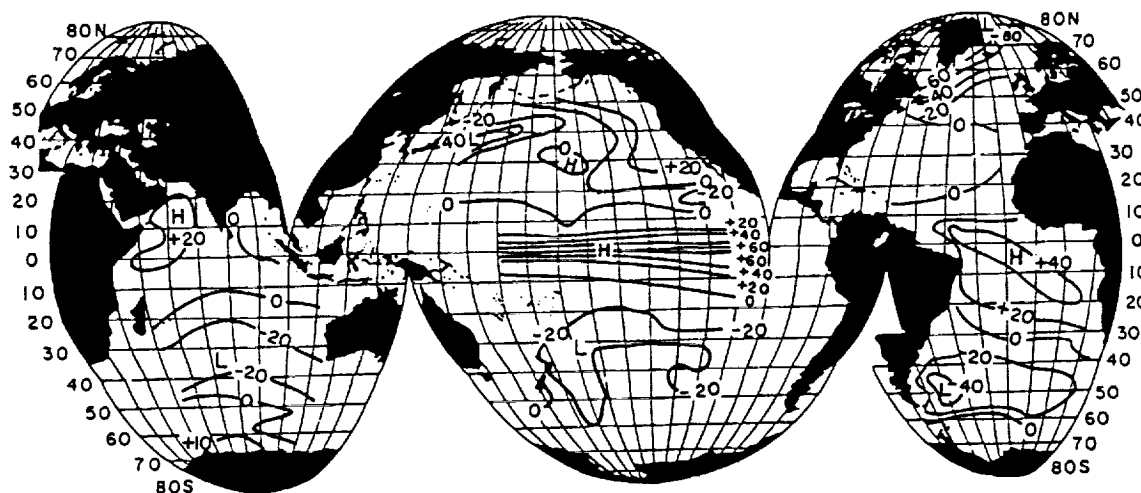
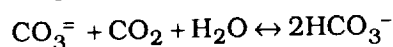


Figure 1. Map showing the difference in  $p\text{CO}_2$  (in  $\mu\text{atm}$ ) between the surface ocean and the atmosphere. Positive values indicate that the ocean is a source of  $\text{CO}_2$  to the atmosphere; negative values indicate an oceanic sink. Based on data collected during the GEOSECS expedition (from Broecker and Takahashi, 1984).

### The Buffer Factor

While  $\text{CO}_2$  in the atmosphere is nonreactive,  $\text{CO}_2$  in the ocean is involved in a variety of inorganic and organic reactions. DIC is made up of bicarbonate and carbonate ions, as well as  $\text{CO}_2$ . The three pools are related through the reversible reaction:



Most of the dissolved carbon in sea water is in the form of bicarbonate ion (90%), with the remainder made up primarily of carbonate ion. Less than 1% of the total DIC in the sea occurs as dissolved  $\text{CO}_2$  gas. As a result, the carbon system in the ocean is well buffered. The strength of this buffering is described by the Revelle factor, which is about 10 for most of the world ocean. A 10% increase in atmospheric  $\text{CO}_2$  requires only a 1% increase in oceanic  $p\text{CO}_2$  to restore equilibrium.

### The Biological Pump

The partial pressure of  $\text{CO}_2$  in the surface ocean is also influenced by biological productivity. Phytoplankton in the sunlit surface layers of the ocean take up  $\text{CO}_2$  and nutrients, and release  $\text{O}_2$ , in forming organic matter by photosynthesis. Most of the primary production is quickly recycled through plant and animal respiration in the upper ocean, replacing the carbon removed by photosynthesis. The "regenerated" production thus does not result in a net removal of  $\text{CO}_2$  from the euphotic zone. The fraction of the primary production that sinks

out of the euphotic zone is known as the new production. If the carbon and nutrient distributions are in steady state, then the carbon and nutrients exported as new production must be replaced by the mixing up of carbon- and nutrient-rich water from below.

Because nutrient concentrations are close to zero in most of the mid- and low-latitude surface ocean, the supply of nutrients has traditionally been considered the limiting factor for phytoplankton growth. In the equatorial Pacific and at high latitudes, however, there are measurable levels of nitrate and phosphate in the surface ocean. The limiting factor in these regions has been assumed to be light limitation. Recently Martin and co-workers (Martin and Fitzwater, 1988; Martin and Gordon, 1988; Martin et al., 1990) have suggested that iron is the limiting factor in many areas far from land (see Peng et al., this volume, for a discussion on limits to the proposed effect of iron).

Nitrogen, phosphorus, oxygen, and carbon are believed to occur in marine organic matter in fixed ratios known as the Redfield ratios (Redfield et al., 1963). In a nutrient-limited system typical of most of the ocean, the net carbon export, or new production, can thus be equivalently expressed in terms of a net input of nitrate.

As the particulate matter exported from the euphotic zone sinks through the water column, it dissolves or decomposes, consuming oxygen and releasing carbon. The downward transfer of carbon from the sea surface to the deep water is known as the biological pump. In the present-day ocean, the biological pump is responsible for maintaining deep ocean concentrations of dissolved inorganic carbon at levels 10–15% higher than surface values (Broecker and Peng, 1982). Because nutrients are also taken up in the formation of organic matter and regenerated at depth, the concentrations of nitrate and phosphate (as well as other minor nutrients needed for plant growth) increase with depth in a way similar to DIC. However, the maximum DIC concentration lies deeper in the water column than the nutrient maxima. This reflects the dissolution of calcium carbonate tests, which occurs at greater depths than the regeneration of organic soft tissue. The sinking and dissolution of calcium carbonate particles carries carbon but not nutrients to the deep sea. The difference in carbon regeneration depths resulting from the “soft-tissue pump” and the “carbonate pump” (Volk and Hoffert, 1985) will be seen to provide an essential mechanism by which biological productivity can influence atmospheric  $p\text{CO}_2$ .

Recent observations of larger concentrations of dissolved organic matter (DOM) than previously assumed have thrown the traditional one-dimensional view of the biological pump, as sketched above, into doubt. In particular, the existence of a large reservoir of non-sinking organic matter with an intermediate lifetime (longer than

days and shorter than centuries) implies that lateral transport may be an essential element of marine biogeochemical cycles (Sugimura and Suzuki, 1988; Toggweiler, 1989).

### The Role of Ocean Transport

As mentioned in the introduction, the ocean plays a direct role in the climate system through the storage and transport of heat. The ocean circulation also plays a more indirect role in the climate system through its impact on the carbon cycle. Once  $\text{CO}_2$  has reached equilibrium between the atmosphere and ocean, there are two main pathways for carbon between the surface ocean and the interior: a biologically mediated path (the biological pump described above) and a path mediated by physical processes. The biological pump is, of course, itself controlled by physical processes (Figure 2); since the new production is, by definition, equal to the rate of supply of nutrients to the upper ocean, the rate of vertical exchange between deep and surface water is the rate-determining step in the biologically mediated uptake of carbon.

The wide variety of physical mechanisms that lead to an exchange between surface and subsurface waters also contribute to the net

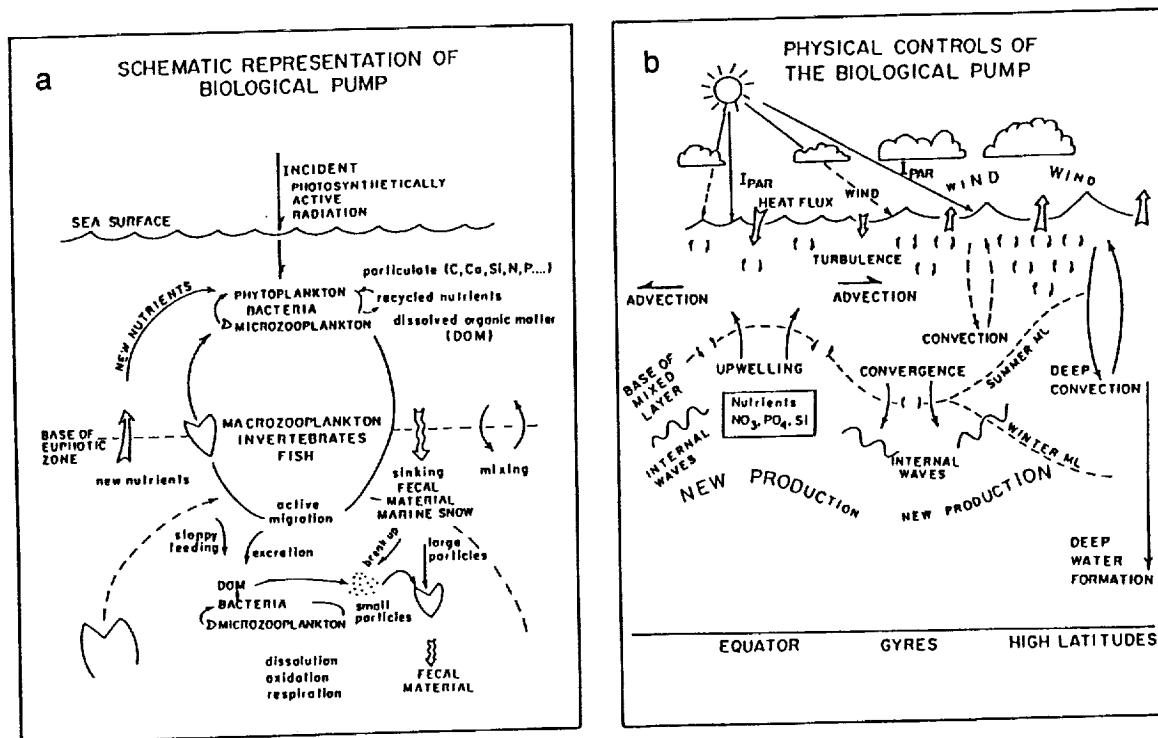


Figure 2. Schematic representation of (a) the biological pump and (b) the physical processes controlling the biological pump (from Bishop, 1989).

exchange of carbon with the ocean interior. Each of these ventilation mechanisms (e.g., subduction, Ekman pumping, the formation of deep and intermediate water masses) has its own characteristic overturning time, which determines the period for which carbon carried into the interior is sequestered from contact with the atmosphere. The physical processes that need to be resolved in a coupled biological-physical climate model therefore depend on the time scale of interest.

As a consequence of the temperature dependence of the solubility of  $\text{CO}_2$ , the thermohaline circulation plays a role in the redistribution of carbon analogous to the role it plays in the global heat balance. Surface water cooled at high latitudes picks up  $\text{CO}_2$ ; when this water loses sufficient buoyancy to sink from the surface layer to the ocean interior, it carries  $\text{CO}_2$  with it. The  $\text{CO}_2$  sequestered by the formation of deep water masses is ultimately given back to the atmosphere through upwelling at low latitudes. Volk and Hoffert (1985) have termed this mechanism of downward carbon transport the "solubility pump."

Another example of how ocean physics and biology interact in the transport of carbon out of the surface layer involves seasonal variability. Codispoti et al. (1982) have shown that in the Bering Sea, biological productivity can draw down surface  $\text{pCO}_2$  by 200 ppmv in two months (Figure 3). Ocean ventilation processes also vary on seasonal time scales. Consequently, biological processes can play an important role in determining the carbon concentration of surface waters carried into the deep ocean by physical mechanisms (Brewer, 1986). Few carbon cycle models presently represent the seasonal cycle of the physics or biology.

### Summary

To sequester atmospheric  $\text{CO}_2$  in the ocean interior requires, first, exchange across the air-sea interface and, second, transfer from the sea surface to the deep ocean. The second transfer is achieved by one of two primary mechanisms: a biologically-mediated path known as the biological pump and a variety of physical processes that result in exchange of water between the upper and lower layers of the sea.

For the purpose of constructing coupled physical-biogeochemical carbon cycle models, it is necessary to look more closely at the relative importance of each of these barriers to the uptake of atmospheric  $\text{CO}_2$  by the ocean. First, concerning gas exchange, the surface ocean and the atmosphere equilibrate with respect to  $\text{CO}_2$  in about one year (Broecker and Peng, 1982), so that gas exchange is not the rate limiting step in the oceanic uptake of  $\text{CO}_2$ . Rather, the rate of vertical exchange between the surface and deep ocean appears to limit the rate at which  $\text{CO}_2$  is taken up by the ocean (Oeschger et al., 1975; Sarmiento et al., 1990). This suggests that

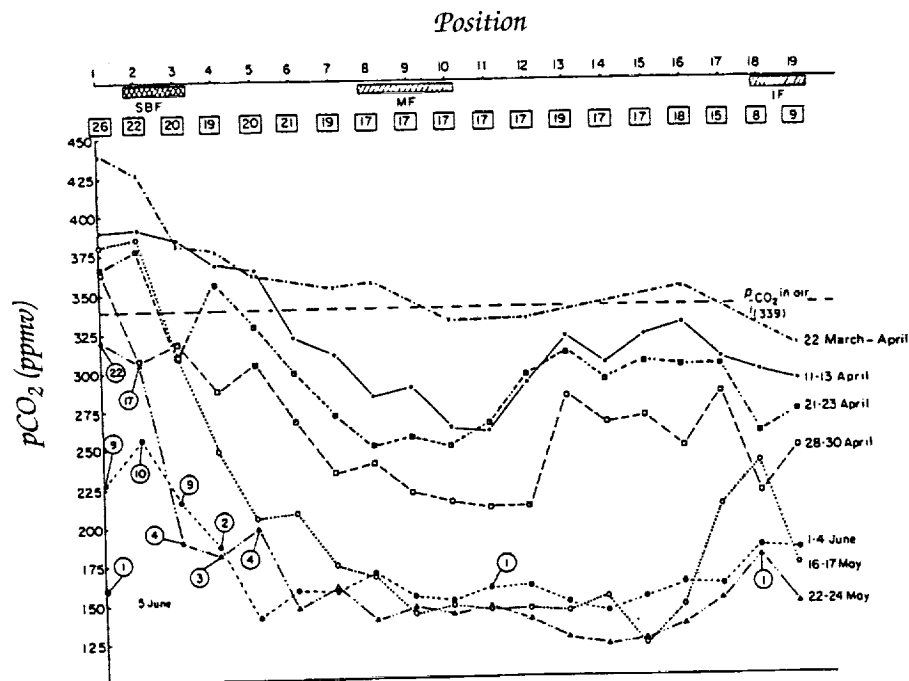


Figure 3. The temporal evolution of  $p\text{CO}_2$  in the surface water along a transect in the Bering Sea. The hatched boxes mark the location of three fronts. Initial nitrate concentrations (in  $\mu\text{g-atom/L}$ ) at the surface are given in the squares. Surface nitrate concentrations  $>1$  between May 22 and June 5 are shown in circles (from Codispoti et al., 1982).

the remaining uncertainties in the gas transfer coefficient are not critical for the prediction of  $\text{CO}_2$  uptake. Supporting evidence for this conclusion is provided by some of the models discussed below, in which the uptake of carbon is found to be insensitive to changes in the gas exchange rate.

The second issue that it is important to make clear concerns the efficiency of the biological pump in the sequestering of carbon. The ocean biota undoubtedly play a critical role in determining the preindustrial, unperturbed, distribution of carbon in the sea. As mentioned above, the downward flux of carbon fixed by marine plants and repackaged by heterotrophs maintains deep ocean concentrations of DIC at levels 10–15% higher than those in the surface ocean. As a result, atmospheric  $\text{CO}_2$  concentrations are significantly lower than they would be if the atmosphere overlay an abiotic ocean.

However, the fact that the marine biota play a crucial role in the natural, unperturbed carbon cycle does not automatically imply that it is essential to include biology in climate models designed to



predict the response to changing external forcing, such as the anthropogenic input of  $\text{CO}_2$ . The models discussed in the following section show that the question of whether biology is important depends on the time scale of interest. The marine biota have little *direct* effect on the ocean's uptake of *anthropogenic*  $\text{CO}_2$ . Primary production in the sea is limited by light and nutrients, not carbon. Increasing the  $\text{CO}_2$  concentrations in the surface ocean therefore has no direct effect on primary production. It is also important to realize that *in equilibrium* the biological pump plays *no* role in the sequestering of carbon in the deep ocean. By definition, if the system is in steady state, the biologically mediated export of organic carbon from the surface ocean is compensated for by an input of inorganic carbon supplied by vertical mixing.

Ocean biology may have an important indirect effect on oceanic uptake of the "excess"  $\text{CO}_2$  supplied by human activities. Changes in the atmospheric circulation due to  $\text{CO}_2$ -induced changes in the radiative forcing will force changes in the ocean circulation. Changes in the circulation, for example, a change in the strength of the thermohaline overturning, may lead to a redistribution of nutrients which will affect the marine biology.

The important point is that changes in the ocean biogeochemical system of relevance to the uptake of carbon from the atmosphere can occur only on relatively long time scales. A global redistribution of nutrients, for example, requires a time equal to the turnover time of the deep ocean, about 1000 years. Therefore, the importance of the role played by marine biogeochemistry depends on the time scale considered; for the prediction of oceanic uptake of carbon on decadal time scales, physical exchange processes are probably most important, while on time scales of 100–1000 years and longer the role of marine biota may be critical.

## **Gaps in Present Physical and Biogeochemical Models**

### **Gaps in the Physical Models**

Before considering particular examples of coupled biogeochemical-physical models, it is worth considering in a general way the characteristics of the physical and biological models that are to be coupled. Most models of ocean biogeochemistry start with a conservation equation for some biologically active component (e.g., phytoplankton or nitrate). The conservation equation includes terms for advection and diffusion, both lateral and vertical, and a biological source/sink term. The first step, then, is an accurate description of the transport field in the ocean. Are there gaps in our knowledge of ocean dynamics?

Modelers of ocean dynamics have an immediate advantage over biogeochemical modelers: The equations are known. We are confident that the Navier-Stokes equations describe the physics of ocean flow accurately. However, it is not possible to resolve every scale of motion, from basin scale to molecular scale. The art of the modeler lies in the parameterization of subgrid-scale processes in a way that maintains the essential nature of the unresolved processes while simplifying the problem enough to make it tractable.

In ocean models, small-scale processes are usually parameterized by relying on an analog to Fick's law of diffusion (Bryan, 1979). The net effect of small-scale mixing is expressed as an eddy coefficient multiplied by the large-scale property gradient. Mixing in the ocean is the result of a large variety of processes acting on different scales (e.g., breaking internal waves, shear instability, convection, double diffusion, wind stirring). It is not clear that the same parameterization should apply to the net effect of each of these processes. To make matters worse, the magnitude of the diffusion coefficient is generally chosen for numerical stability, rather than on physical grounds; the coarser the resolution, the larger the diffusion coefficients must be. Consequently, in coarse-resolution models the thermocline tends to be too diffuse, fronts tend to be too broad, and inertial effects are too strongly damped.

The results of a model are generally sensitive to the magnitude of the diffusion coefficient. Bryan (1987), for example, has demonstrated the sensitivity of the thermohaline circulation and meridional heat flux to the magnitude of the vertical diffusivity in a coarse-resolution general circulation model (GCM). Given the importance of the thermohaline circulation in the climate problem, as discussed above, such sensitivity is cause for concern.

As the model grid size decreases and more scales of motion are explicitly resolved, the relative magnitudes of "advection" and "diffusion" change. Eddy resolving GCMs do not have to parameterize as much of the eddy field and can use smaller eddy coefficients. However, such calculations are extremely computationally intensive, and it is not possible to integrate long enough for the thermohaline circulation to reach equilibrium or to do multiple experiments to test the sensitivity to the model parameterizations.

A final issue related to the parameterization of mixing in ocean circulation models regards the orientation of the mixing tensor. Mixing in the ocean occurs much more efficiently along neutral surfaces than across them, since motions across neutral surfaces are opposed by buoyant restoring forces. In most ocean models, lateral mixing is represented as a strictly horizontal process. Since neutral surfaces are generally inclined to the horizontal, mixing in the

model does not truly represent the effect of turbulent motions on the transport of physical and chemical properties. Lin (1988), for example, has shown that heat penetrates into the ocean interior more efficiently in a model with isopycnal diffusion than in one with horizontal diffusion. Similarly, the injection of chemical properties such as  $\text{CO}_2$  will be sensitive to the orientation of the mixing tensor.

A second major problem encountered by an ocean modeler is the fact that the forcing at the ocean boundary (wind stress, heat flux, and freshwater flux) is usually poorly known. Wind stress and precipitation are notoriously difficult to measure accurately from ships. The heat flux and evaporation rate also depend critically on the wind speed. Generally the fluxes have been so poorly known that modelers have resorted to tricks such as restoring the surface temperature and salinity to "observed" values with some time constant. There are several potential problems with this approach: The "climatological" average values used as observations generally are highly smoothed and may not reflect the spatial/temporal scales of the real surface fluxes; the time constant introduces an additional free parameter; and there is the possibility that one can mask some fundamental deficiency in the model physics by preventing the model solution from drifting too far from the "truth." Perhaps the main problem is that we cannot use such a model to predict the response to changing conditions in the future, for which observations are not available, or to couple to an atmospheric model.

### **Gaps in the Biological Models**

If our goal is the design of a coupled model that is able to predict the carbon export from the euphotic zone, what kind of biological model is required, and where are the gaps in present models?

Given that  $\text{CO}_2$  is not limiting in the surface ocean environment, the two fundamental requirements for primary production are light and nutrients. At a minimum, then, a coupled biological-physical model for studies of the oceanic carbon cycle must reproduce the light and nutrient environment experienced by primary producers. The light and nutrient regimes are functions of the vigor of vertical mixing in the upper ocean. If the vertical mixing is very weak, the absence of an input of nutrients will limit primary production; if the vertical mixing is deep and energetic, phytoplankton spend only a short time in the euphotic zone, and production is again limited.

In addition, the structure of marine communities, or the relative abundance of different species, changes as the vertical mixing regime changes to take maximum advantage of the resources available. For example, stably stratified regimes resulting from weak vertical mixing tend to favor small phytoplankton, while energetic

regimes characterized by episodic pulses of nutrients tend to be dominated by large phytoplankton with rapid growth rates (Williams and von Bodungen, 1989). The former community type efficiently recycles the limited supply of nutrients, and only a small fraction of the total primary production is exported. In contrast, the communities dominated by large cells such as diatoms export a relatively large fraction of the primary production. Therefore changes in the physical mixing regime can lead to changes in community structure and the efficiency with which biological processes export carbon from the euphotic zone.

Note, however, that while a shift from low mixing to high mixing at a particular location may lead to shifts in community structure and changes in the new production, the net uptake of carbon is unchanged; the supply of "new" nutrients by mixing will be accompanied by a supply of DIC in the Redfield proportion. The only way this constraint can be removed is if the ratios of C:N:P are not constant in marine organic matter. Although observations suggest that the Redfield ratios are relatively constant, the extent to which the Redfield ratios may vary spatially, temporally and by species is not well known.

An accurate description of the time history of the nutrient distribution and the mixed-layer depth would therefore appear to be essential for a coupled biological-physical model. An additional step is necessary to translate the supply of nutrients and light into a net export of carbon. A variety of "rules" have been used to relate the export production to the physical mixing, light, and nutrient distributions.

The simplest "rule" is to simply equate the export production to the input of "new" nutrients by mixing, using the Redfield ratio to convert the flux of nitrate or phosphate into an equivalent carbon flux. The majority of the models discussed in the next section adopt this approach. It has the advantage of simplicity and serves as an integral constraint that may be applied without a knowledge of the details of the biological system. However, there are several limitations to this approach. First, the method implicitly assumes that nutrients are the sole factor limiting primary production. This is not true at high latitudes, or even at midlatitudes during the winter, when light is limiting. Second, the method assumes that the Redfield ratio is constant and accurately known. Third, it does not take into account some evidence that other nutrients, including silicate, and elements needed in trace quantities, such as iron, may be limiting in some regions at some times of year. Finally, the method is of no use if we are interested in issues other than the net export production, such as the standing stock of phytoplankton or the dynamics of the spring bloom.

At the opposite extreme lie ecosystem models that attempt to explicitly resolve the different classes of organisms, and the transfers between them, that make up the marine food web. The immense complexity of the biological system is usually reduced to a set of pools of carbon or nitrogen representing the components of the food web (e.g., phytoplankton, zooplankton, bacteria, nitrate) (Fasham, 1985; Venzina and Platt, 1987; Michaels and Silver, 1988). Each of these pools can be advected and diffused. In addition, biochemical processes transfer material from one pool to another, but the equations describing these transfers are not well known. A large number of parameters, which generally are poorly constrained by observations, are needed to describe such factors as the efficiency of carbon transfer between trophic levels, the sinking and regeneration rate of particles, and the physiological characteristics of individual cells (e.g., maximum growth rate, photosynthetic response to light). Plankton adapt to changing environmental conditions, so that these parameters may also change in time depending on the past history of the cell.

The minimum ingredients necessary to model a food web are light, nutrients, a primary producer (to fix carbon), and a secondary producer (to package the fixed carbon for export). Additional elements that may be critical to model a particular phenomenon include multiple size classes of phytoplankton and zooplankton, multiple phytoplankton species (e.g., diatoms forming silicic acid tests vs. coccolithophores forming calcium carbonate tests), ammonium as well as nitrate as a substrate, trace nutrients such as iron, vertical migration, particle aggregation to form "marine snow," cannibalism, higher trophic levels, and production of dissolved organic matter. To include these additional pools or processes, the transfer of carbon between each of the pools must be parameterized. However, the rate of transfer between various reservoirs is extremely difficult to observe, and the parameterizations are thus highly uncertain.

Therefore, one category of gaps in present biological models is rooted in the question of how to parameterize the many interactions in a complex biological system that is patchy in space and time. More fundamentally, we do not have a clear understanding of which of these processes are essential and which can be safely ignored. We have even less feeling for the importance of resolving the nonlinear interactions occurring between physical and biological processes that are varying on similar spatial and temporal scales.

To consider a specific example, the box models discussed in the next section suggest that changes in biological productivity at high latitudes, where surface nutrient concentrations are not equal to zero, can have a profound impact on atmospheric  $p\text{CO}_2$  on the time scale of centuries. One question we might want to ask of our cou-

pled ocean circulation-carbon cycle model is this: Given the changes in wind stress and heat flux expected at the sea surface due to CO<sub>2</sub>-induced changes in the radiative forcing of the atmosphere over the next 100 years, how will high-latitude circulation, productivity, and distribution of nutrients and carbon change? (Ultimately, of course, the atmospheric model would be coupled directly to the ocean carbon model). At the present time, we have no idea what elements of the marine food web—or, for that matter, of the physical model—need to be included to answer this question.

### Summary

Basin-scale ocean circulation models have generally focused on the large-scale flow. The limited spatial resolution achievable in these models has meant that several aspects of the ocean that are potentially the most important in determining the role of the ocean in the global carbon cycle have been neglected. In particular, the dynamics of the mixed layer and exchange between the mixed layer and the ocean interior are not well reproduced in present coarse-resolution models. Deficiencies in the way present models simulate ocean ventilation processes, together with uncertainties in the surface forcing, make model-based predictions of the oceanic uptake of carbon by physical processes highly uncertain.

The difficulties are enhanced when one contemplates coupling the physical and biogeochemical subsystems in a model. The biogeochemistry is most sensitive to exactly those aspects of the ocean that coarse-resolution models have the most trouble reproducing: the physical and chemical properties of the mixed layer and vertical exchange processes. Accurate representation of upper-ocean mixing and knowledge of the surface forcing are critical to simulate both the physically and biologically mediated transfer of carbon to the ocean interior.

It is clear that the circulation will depend on the resolution of the model, but we do not understand the biogeochemical processes well enough to know what features of the circulation we need to get right. For example, even the eddy resolving models do not accurately reproduce some basic features, such as the location of the separation of the Gulf Stream from the coast. The Gulf Stream plays a major role in determining the pattern of mixed layer depth in the North Atlantic; is getting the Gulf Stream separation right therefore important for modeling the carbon cycle? How much trust can we have in the biogeochemical implications of a circulation model that has clear physical inadequacies? The remainder of this paper reviews a variety of attempts to model aspects of marine biogeochemical cycles in the search for answers to these questions.

## Examples of Coupled Ocean Dynamics-Biogeochemical Models

### Box Models

The use of box models to study the carbon cycle has a long history (e.g., Revelle and Suess, 1957; Craig, 1957; Nydal, 1968; Oeschger et al., 1975). The box-diffusion model of Oeschger et al., in which the ocean is represented as a deep diffusive reservoir capped by a well-mixed surface box in communication with the atmosphere, has proved particularly useful. Sixteen years after its introduction, the box-diffusion model (with some modifications) is still the primary tool used to predict the carbon cycle's response to different CO<sub>2</sub> scenarios (e.g., Houghton et al., 1990).

The analysis of trapped air bubbles in ice cores, which showed that atmospheric pCO<sub>2</sub> increased at the end of the last glacial period from 200 ppm to 280 ppm in only a few hundred years (Neftel et al., 1982), sparked a renaissance in biogeochemical box modeling. This observation demonstrated that the carbon system was capable of much more rapid change than had been previously thought. A series of simple box models was developed to explore the role that the oceans might play in producing such rapid changes in atmospheric pCO<sub>2</sub> (Broecker, 1982; Sarmiento and Toggweiler, 1984; Knox and McElroy, 1984; Siegenthaler and Wenk, 1984). These models provide some of the clearest suggestions of how ocean circulation and biology may interact to determine atmospheric pCO<sub>2</sub>.

In box models the ocean is reduced to a small number of well-mixed reservoirs, and the net effect of all ocean transport processes is parameterized as a few exchange terms between reservoirs. The magnitude of the "circulation" is fixed by fitting to the basin-averaged vertical profile of a tracer such as <sup>14</sup>C. Biology also enters in the simplest possible way: The export from the surface box is proportional to the nutrients supplied by mixing between the surface box and the nutrient-rich deep box.

Broecker (1982) used a two-box model of oceanic phosphate to study the increase in atmospheric pCO<sub>2</sub> that occurred at the end of the ice age. In this model, the "physics" consists of a single term, the exchange between the boxes. Broecker proposed that during the glacial period, phosphate concentrations increased due to oxidation of organic matter deposited on the continental shelves. Increased oceanic phosphate led to an increased efficiency of the biological pump and a larger transport of carbon to the deep sea. More recent work, however, has shown that the changes in atmospheric pCO<sub>2</sub> preceded the changes in ice volume (Shackleton and Pisias, 1985).

and occurred too rapidly (Siegenthaler and Wenk, 1984) to be consistent with this hypothesis.

The next generation of box models overcame these difficulties by dividing the surface ocean into a high- and a low-latitude box (Knox and McElroy, 1984; Sarmiento and Toggweiler, 1984; Siegenthaler and Wenk, 1984). Phosphate at high latitude ( $P_{hl}$ ) is determined by the balance between the input of nutrient-rich water from below and the export of nutrients in sinking organic matter. When convective exchange between the surface and deep boxes at high latitude increases,  $P_{hl}$  increases, which in turn leads to increases in surface DIC and alkalinity, and thus atmospheric  $pCO_2$ . Increases in high-latitude productivity export more nutrients to the deep sea,  $P_{hl}$  decreases, and hence so does atmospheric  $pCO_2$ . The presence of a high-latitude outcrop allows more direct communication between the atmosphere and the carbon reservoir in the deep ocean. This in turn permits relatively rapid exchange between the atmospheric and oceanic carbon reservoirs.

A somewhat different three-box model was used by Dymond and Lyle (1985) and Sarmiento et al. (1988) to investigate the effect of differences in the regeneration depth of organic carbon and calcium carbonate on atmospheric  $pCO_2$ . The simplest model they could devise that still contained the essential features consisted of three stacked boxes, one each for surface, thermocline, and deep water. All organic matter was assumed to be regenerated in the thermocline box, and all calcium carbonate was assumed to regenerate in the deep water box. In this model, atmospheric  $pCO_2$  is changed by sequestering or releasing ALK in the deep ocean; an increase in the sequestering of ALK in the deep sea or an increase in the surface mixing will lead to an increase in atmospheric  $pCO_2$ .

Recently the number of boxes included in such models has continued to grow (e.g., Broecker and Peng, 1986; Keir, 1988; Boyle, 1988; Volk, 1989). The models with additional boxes are constructed in the same way as the two- and three-box models: The circulation is postulated a priori, and the resulting atmospheric  $pCO_2$  and carbon and nutrient distributions are calculated. However, additional boxes do reveal phenomena that are hidden in the simpler models. For example, Keir (1988) stressed the essential difference between the Mediterranean-type high-latitude circulations typical of the North Atlantic (and of the three-box models above) and the estuarine-type circulations thought to be characteristic of the Antarctic. In the first case, changing the productivity at high latitudes decreases atmospheric  $pCO_2$  but has little effect on low-latitude productivity, since the nutrient concentration of deep water upwelling into the low-latitude surface box has not changed signifi-



cantly. In the second case, increasing productivity in the Antarctic leads to trapping of carbon and nutrients in the circumpolar deep water. This decreases both atmospheric  $p\text{CO}_2$  and productivity in the warm surface waters, since the preformed nutrient concentration of the intermediate water upwelling to supply the warm surface box is decreased.

The primary justification for box models is that they provide insight into the essential mechanisms controlling  $p\text{CO}_2$  and help to identify those parameters to which the properties of interest are most sensitive. For example, the box models have indicated that high-latitude production, the "alkalinity pump," and the magnitude of vertical mixing all may play an important role in determining atmospheric  $p\text{CO}_2$ . However, the results of box models should be viewed with caution; even in these simple systems there may be multiple scenarios consistent with a particular observation (such as the fact that atmospheric  $p\text{CO}_2$  was 80 ppm lower in glacial times). The comparison of different box models also demonstrates the sensitivity of the results to the model configuration and "physics" assumed, such as the magnitude and sense of the overturning circulation. For example, the two-box model suggested that ocean biology played a small role in determining atmospheric  $p\text{CO}_2$ ; dividing the surface ocean into a high- and a low-latitude box, as in the three-box models, suggested instead that biological productivity at high latitude could have a major impact (Sarmiento et al., 1988). Most of the present box models have been calibrated with a single tracer, usually natural  $^{14}\text{C}$ . Models calibrated with other tracers such as bomb-produced  $^{14}\text{C}$  or tritium result in different values for the mixing coefficients and circulation (Sundquist, 1985). Furthermore, none of these tracers is a perfect analog for  $\text{CO}_2$ . Finally, such models cannot tell us how the system might change. This suggests the need for models with less extreme parameterization.

### Diagnostic Box Models

Diagnostic or inverse models differ from box models primarily in that some ocean dynamics is included. Diagnostic models attempt to deduce the circulation, mixing, and biochemical transformation rates consistent with a set of physical, biological or chemical observational constraints (e.g., Riley, 1951; Garcon and Minster, 1988; Schlitzer, 1988; Bolin et al., 1989; Metzl et al., 1990; Rintoul and Wunsch, 1991). The inverse calculations share some of the advantages and disadvantages of box models: Such models are often useful for identifying important physical or biochemical mechanisms, but are of little use in determining how the system may change in the future.

Furthermore, like those of box models, the conclusions reached by these models may depend on the degree of resolution or averaging of the data. For example, Schlitzer (1988) concluded from a low-resolution model of the North Atlantic that dissolved organic matter played no role in the basin-scale nitrogen balance. Schlitzer avoided the Gulf Stream, realizing that the highly smoothed data he used would not adequately represent the sharp front of the western boundary current. Rintoul and Wunsch (1991), on the other hand, considered a smaller area but with high-resolution hydrographic data. They found that the western boundary current played a critical role in the transport of nitrate and that a sizeable dissolved organic nitrogen (DON) pool was required to balance the nitrogen budget in the North Atlantic. When the data were smoothed prior to the inversion, the meridional fluxes of nitrate and silicate were changed significantly.

Although inverse models are by nature not predictive tools, once a solution has been found, it can be used to perform transient or predictive calculations by assuming the circulation and mixing remain in steady state. For example, Bolin et al. (1989) estimated the transient uptake of tritium,  $^{14}\text{C}$ , and  $\text{CO}_2$  from the atmosphere by the Atlantic Ocean (Figure 4). They found that large-scale advection was primarily responsible for the transport of carbon into the ocean interior. A second experiment with a more energetic circulation led to greater uptake of carbon.

A great advantage of inverse models lies in the variety of information that can be used to constrain the solution. One also obtains useful information concerning what features of the solution have been well determined by the available information, and explicit error estimates. However, since the primary goal is the development of a predictive coupled model, I will not discuss diagnostic models further.

### Mixed-Layer Models

At the opposite extreme of model scale, models of the mixed layer use high resolution (a few meters) in the upper 100–300 m of the water column and treat the rest of the ocean as a bottom boundary condition. In the open oligotrophic ocean, the input of nutrients into the euphotic zone, and thus new production, is controlled by turbulent vertical transport through the base of the mixed layer. Several attempts have been made in recent years to use models of mixed-layer physics together with some representation of biologically important processes to study the interaction between physics and biology in the upper ocean. Perhaps the most valuable insight that has been gained from these one-dimensional models is an apprecia-

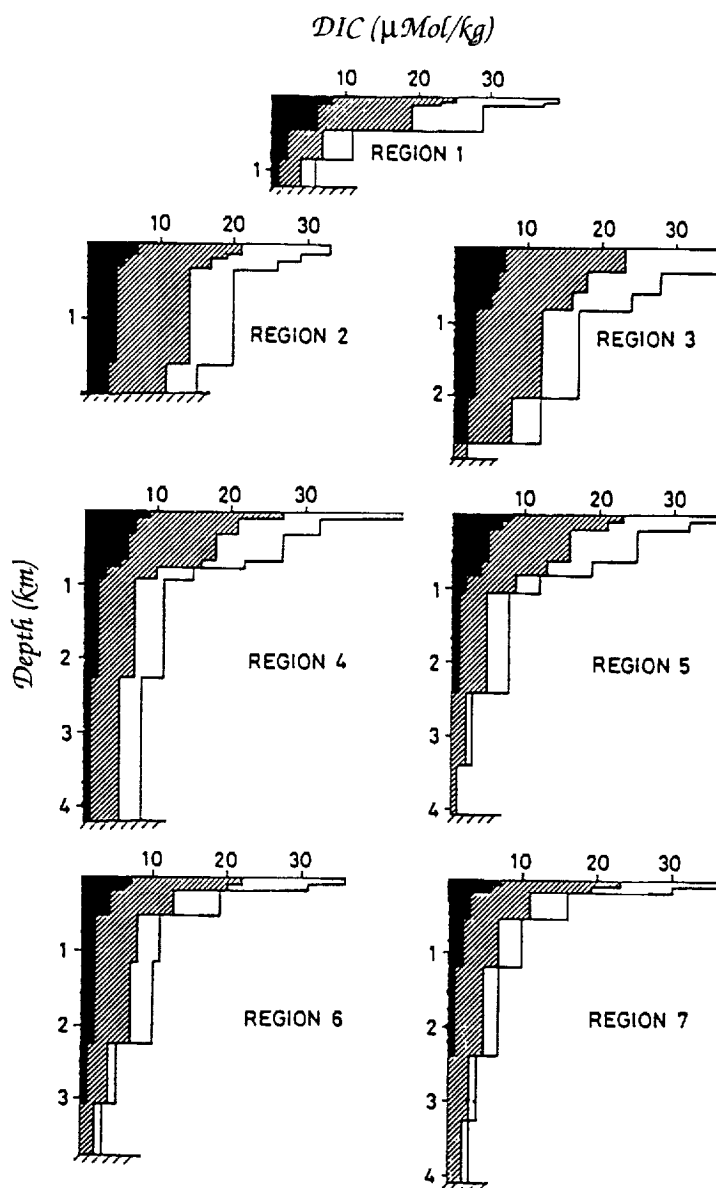


Figure 4. Modeled accumulation of carbon (in the form of DIC) in the Atlantic Ocean as a result of increasing atmospheric  $\text{CO}_2$  concentrations. For selected regions, vertical profiles of  $\text{CO}_2$  concentrations are shown for 1880 (black), 1955 (stippled), and 1983 (white), resulting from a transient calculation using the steady-state circulation. Note the greater concentration in the western basin (regions 2, 4, 6), reflecting transport of carbon into the interior by the formation of North Atlantic deep water (from Bolin et al., 1989).

tion for the importance of nonlinear interactions between physical and biological processes that are varying in time.

Mixed-layer models are generally one-dimensional; lateral advection and mixing are ignored. The depth and properties of the mixed layer depend on atmospheric forcing (wind stress, heat flux, freshwater flux), turbulent vertical mixing, and in situ absorption of solar radiation. The concentrations of dissolved gases, such as oxygen or carbon dioxide, depend on surface gas exchange as well. The mixed-layer models of interest here also include a biologically active component (e.g., phytoplankton or nutrient concentrations). The biological component is passively advected and diffused in the same manner as temperature and salinity, but has additional source/sink terms due to biological or chemical activity.

Kiefer and Kremer (1981) used a mixed-layer model to explain the origin of the subsurface chlorophyll maximum near the base of the euphotic zone. They employed a model of phytoplankton growth based on following flows of nitrogen between three pools: nitrate, nitrite, and phytoplankton. They concluded that the chlorophyll maximum resulted from the formation of the seasonal thermocline by the warming of the surface layers in the late spring. The thermocline stabilized the lower portion of the euphotic zone and isolated the layer in an environment favorable to growth.

While Kiefer and Kremer underscored the importance of *seasonal* changes in the depth of the mixed layer, Woods and Onken (1982) focused on the effect of *diurnal* variation on primary production. The "Lagrangian-ensemble" model of Woods and Onken is very different from the more commonly used "Eulerian-continuum" models, such as that of Kiefer and Kremer. The latter models treat phytoplankton concentration, for example, as a continuum property of the sea water that changes with time at fixed points. A Lagrangian model of phytoplankton, on the other hand, follows the trajectory of individual phytoplankters as they move through a variable environment (e.g., changing light levels or nutrient concentrations). Forming the ensemble average of the individual trajectories has the effect of averaging after integrating the nonlinear equations describing phytoplankton growth. Woods and Onken point out that this is particularly important at diurnal scales, because the physical environment and the physiological responses of the plankton to changing conditions vary on similar time scales.

Woods and Onken were interested in describing the initiation of the spring bloom, and they neglected the effects of zooplankton grazing, nutrient limitation, and self-shading, as well as lateral advection or diffusion, all of which would tend to increase the nonlinearity and further justify the use of Lagrangian methods. How-

ever, even the present simplified system is complex and uses a time step of three minutes for the phytoplankton equations. The prospect of scaling up such a model to large scales is intimidating. On the other hand, both the Kiefer and Kremer and Woods and Onken models result in similar behavior on seasonal time scales, perhaps indicating that we can get away with not resolving the nonlinearities on the scale of individual cells.

Klein and Coste (1984) emphasized a different type of nonlinear interaction in the mixed layer. They were particularly concerned with the effect of variable surface forcing (wind stress) on nutrient transport into the mixed layer. Klein and Coste concluded that the entrainment of nutrients was dominated by the interaction between time-varying wind stress and near-surface currents. In particular, a resonance between the wind stress and wind-generated inertial currents at the inertial frequency (about one day at 30° latitude) resulted in pulses of nutrients entering the mixed layer (Figure 5). This nonlinear interaction affects both instantaneous and mean values of the nutrient flux.

Klein and Coste did not include any active biology within the mixed layer. They focused on the net input of nitrate into the mixed layer, which is (by definition) equal to the new or export production. A different approach to estimating the new production using a mixed layer model has been taken by Musgrave et al. (1988). They have focused on simulating the seasonal cycle of oxygen in the upper ocean. The seasonal buildup of a subsurface oxygen maximum in the subtropical ocean can be used to estimate the amount of new production, since the recycled production loop has no net effect on oxygen concentrations (Shulenberger and Reid, 1981; Jenkins and Goldman, 1985). Moreover, since the oxygen produced during photosynthesis remains after the "new" organic matter has sunk out of the mixed layer, the oxygen concentration tends to average over a series of episodic production pulses (resulting, for example, from the pulses of nitrate input described by Klein and Coste).

The physical model employed by Musgrave et al. is based on that of Price et al. (1986). The equations include terms expressing the dependence of temperature, salinity, and horizontal velocity on entrainment of water from below; surface fluxes of heat, salt, and momentum; vertical advection; vertical diffusion; in situ absorption of solar radiation; and inertial rotation of the horizontal velocity due to the Coriolis force.

The rate of change of mixed-layer depth is needed to close the set of equations. Musgrave et al. express the entrainment rate as a function of convection forced by surface cooling and evaporation, wind mixing, and shear instability at the base of the mixed layer.

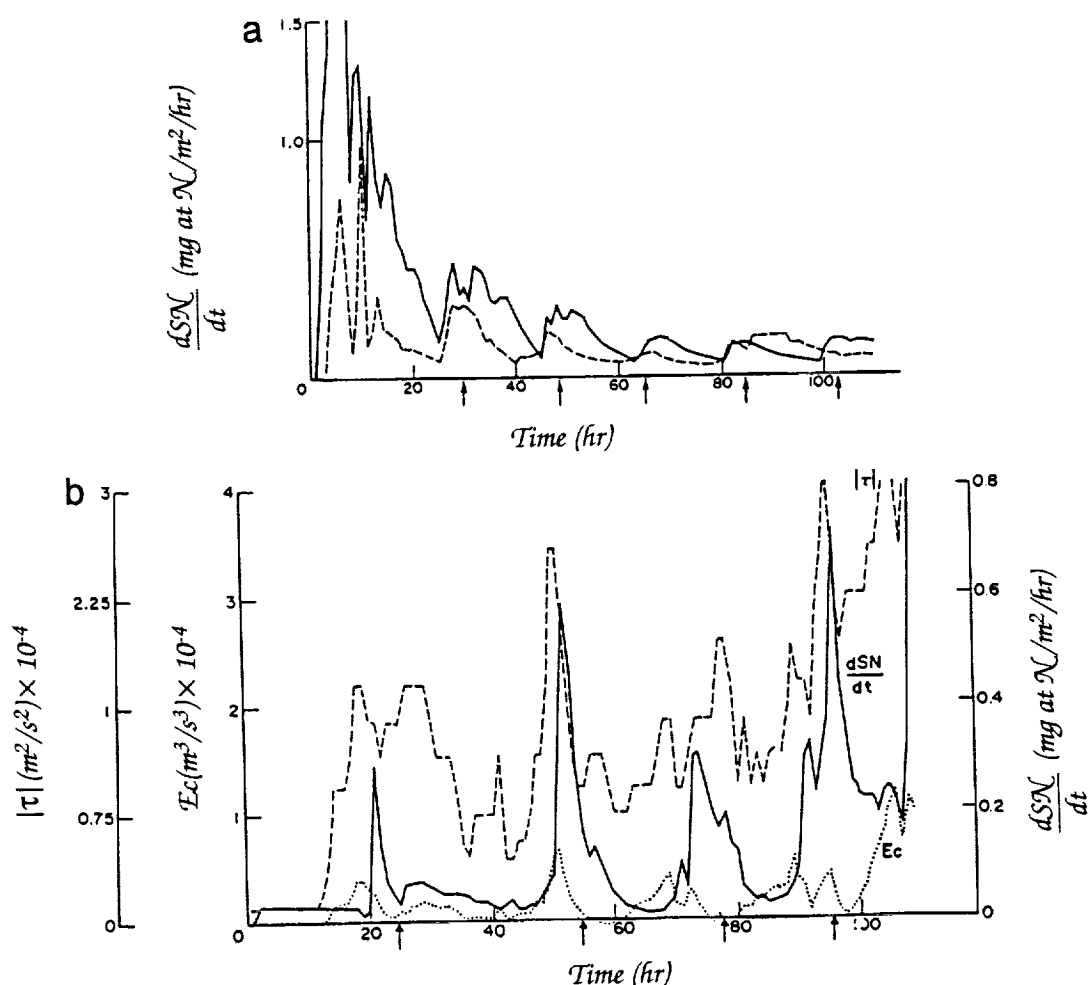


Figure 5. (a) Time variation of the nitrate flux ( $dSN/dt$ ) into the mixed layer in response to constant wind stress. Solid line: wind stress  $= 4 \times 10^{-4} m^2/s^2$ ; dashed line:  $2 \times 10^{-4} m^2/s^2$ . (b) Time variation of nitrate flux ( $dSN/dt$ , solid) in the case of varying wind stress ( $|\tau|$ , dashed).  $Ec$  (dotted) is the energy supplied to the sea surface by the wind stress (from Klein and Coste, 1984).

The model thus requires specified surface fluxes (wind stress, heat, fresh water) as well as boundary conditions at 300 m for temperature, salinity, diatomic oxygen, and velocity. Monthly averages were used for the surface fluxes other than wind stress. Wind stress is proportional to the square of the wind speed, so high winds, which are not properly represented by monthly averages, can contribute significantly to the surface momentum budget. To simulate the effect of periodic storms, Musgrave et al. superimposed a stochastic wind spectrum on the monthly averaged wind speeds.

Musgrave et al. reached the following conclusions:

- The resonant interaction between the wind stress and inertial currents noted by Klein and Coste (1984) was a crucial part of the model physics, implying that resolving the inertial period was important.
- Using the thermal cycle for calibration limited the vertical mixing coefficient to a fairly narrow range.
- The vertical mixing was still sensitive to the surface fluxes and the parameterization of the penetration of solar radiation. Therefore, the optical properties of sea water at a given location (which may change in time) must be known to determine the vertical mixing there.
- Reproduction of the seasonal oxygen signal required representation of storms using a stochastic wind field.

All of the models discussed above are one-dimensional and neglect lateral processes. Woods and Barkmann (1986) have pointed out that advection of water columns through regions of varying surface forcing may have a large impact on mixed-layer properties, and hence primary production. Recent measurements of a larger-than-expected pool of dissolved organic matter also suggest that lateral transport may be a critical link in basin-scale carbon and nutrient cycles.

A further example of the importance of spatial variations in determining the net export from the euphotic zone was given by Nelson et al. (1989). They showed that the shoaling of the main thermocline due to the frictional decay in a Gulf Stream warm core ring led to upwelling in the ring interior. The upwelling did not itself provide an input of nitrate to the euphotic zone; rather, the nitracline shoaled within the ring so that less energetic wind-mixing events were capable of breaking through the nitracline and entraining nutrient-rich water. Thus, lateral variations in the upwelling rate can "warp" the nutricline sufficiently that the nutrient supply is increased, even if the strength of vertical mixing does not change. Moreover, Nelson et al. found that the episodic injection of nitrate led to elevated levels of nitrate uptake by the phytoplankton, so that the ratio of new to regenerated production (the  $f$  ratio) was very high (about 0.6). The global significance of locally enhanced new production in mesoscale features is unknown.

Each of the models discussed above underscores the importance of nonlinear interactions between physical and biological processes, even with the simple proxies for the biological system used to date. As yet we do not know how important it is to reproduce these inter-

actions if we are to model, for example, basin-scale patterns of new production and their evolution in time. Parameterization in terms of bulk or mean values may be difficult, if not impossible. Progress is likely to come from detailed observational and numerical studies on small scales, which will develop the understanding of the fundamental mechanisms that is the first step in the development of appropriate parameterizations.

### Three-Dimensional Models

Wroblewski et al. (1988) took the next step from one-dimensional toward three-dimensional simulations by considering an  $n \times 1$ -D model, consisting of multiple applications of a one-dimensional mixed-layer model at each point of a  $1^\circ \times 1^\circ$  grid. Using this model, they were able to reveal some of the space and time scales of the spring bloom in the North Atlantic. To include the effect of lateral advection and mixing, however, a three-dimensional circulation model is required.

The first attempts at including a biological model in an oceanic general circulation model have recently begun (Sarmiento et al., 1989; Bacastow and Maier-Reimer, 1990). Sarmiento and co-workers set out to model the seasonal progression of the spring bloom in the North Atlantic. The circulation model is the  $2^\circ \times 2^\circ$  "robust diagnostic" primitive equation model of Sarmiento and Bryan (1982), with increased vertical resolution in the upper ocean. Interior and surface values of temperature and salinity are restored to observed climatological values using a Newtonian damping term.

Embedded within the circulation model is a food web model of upper ocean ecology (Fasham et al., 1990). The food web model follows flows of nitrogen between various pools: phytoplankton, zooplankton, bacteria, nitrate, ammonium, DON, and detritus. Each of these pools is advected and diffused, in addition to the biologically mediated transfers from one pool to another. The ecological model requires estimates of a large number of coefficients that express the efficiency with which organic matter is transferred within trophic levels, the photosynthetic response to light, the relative preference for nitrate or ammonium as a substrate, the fraction of production that enters the DON pool, etc. The parameter values have been selected by calibrating a one-dimensional version of the food web model to a time series of observations at Bermuda.

The time step in the model is limited by the need to resolve the sinking of detritus from the euphotic zone. As a result it is not possible to integrate the model long enough for the deep ocean to reach equilibrium. Currently the model is integrated for three years in each experiment.



The food web model will ultimately include carbon as well as nitrogen, and additional pools such as different size classes of plankton. A model of particle cycling in the water column and a sediment-diagenesis model of  $\text{CaCO}_3$  dissolution and organic carbon oxidation will be incorporated.

Several important lessons have already been learned from the preliminary experiments. First, lateral transport has a major impact on the results, supporting the need for full three-dimensional models. Furthermore, the greatest discrepancies between the model results and the observations are due to the deficiencies of the *physical* model. In particular, the mixed-layer depths are unrealistic in some areas, which has a large impact on the biology, and the equatorial upwelling is poorly reproduced, perhaps due to errors in the wind data. Second, a major constraint is the lack of observations: The only time series measurements of the sizes of the various pools in the food web are from a single site. It is not clear that the parameter values that give the best fit to observations at Bermuda can apply to different oceanic regimes. At the large scale, satellite observations of ocean color provide time series observations to which the model can be compared. However, it is not obvious which model quantity is most appropriate to compare to the satellite measurements. Third, there is some indication that even the present ecosystem model, which is fairly complex, is too simplified to capture the essential dynamics. In particular, additional compartments, including different plankton size classes and detritivores, may be necessary.

Bacastow and Maier-Reimer (1990) have taken a somewhat different approach. The model equations are a filtered version of the full primitive equations, which is appropriate for long period integrations. Using a one-month time step, Bacastow and Maier-Reimer can integrate the model for 5000 years in prognostic mode (i.e., there is no restoring to data to accelerate convergence as in the model of Sarmiento et al.). The biology component of the model is much simpler than that of Sarmiento et al. Rather than model the marine food web explicitly, Bacastow and Maier-Reimer specify the new production as a function of the nutrient concentration in the mixed layer using Michaelis-Menten kinetics. Particles leaving the mixed layer are regenerated in the same grid box. The carbon model is run with a seasonal time step.

Bacastow and Maier-Reimer used their model to explore the differences between the model carbon cycle with and without biota. The patterns of sources and sinks of atmospheric  $\text{CO}_2$  were similar in the two cases, but the magnitudes were changed. In particular, with no biota the difference between DIC concentrations in surface water and deep water was 25% of the surface to deep water contrast

in the model with biology. In the case with no biota, the difference is solely due to the solubility pump. With biology, the carbonate and soft-tissue pumps act to increase the downward flux of carbon. When the circulation is decreased in strength, all three carbon pumps increase in efficiency, decreasing surface water carbon and hence atmospheric  $p\text{CO}_2$ .

### **Exchange with the Continental Margins**

The relative productivity of the coastal ocean vs. the open ocean is the most dramatic feature of satellite maps of ocean color (e.g., Esaias et al., 1986). Upwelling along eastern boundaries and mixing due to energetic eddies near western boundaries provide a source of nutrients supporting the high levels of productivity observed. Higher productivity and a shallower water column near continental boundaries imply a higher probability that organic matter sinking out of the euphotic zone will be buried on the sea floor rather than remineralized within the water column. Several investigators (e.g., Walsh et al., 1981) have therefore concluded that burial of organic matter formed on the continental shelves provides a major oceanic sink of carbon. Rowe et al. (1986), on the other hand, showed that if pelagic microbial consumption and the lag in coupling between seasonal production and consumption were taken into account, then no net export of organic matter from the shelf occurred in the northwest Atlantic. Emerson (1985) has also argued that rates of organic matter degradation are too rapid for significant burial on the continental slope to occur. However, the net flux between the continental margins and the open ocean largely remains an open question. The fluxes are difficult to measure directly, due to the nonlinear interaction of transport and biological processes and the potential importance of rare, but extreme, events such as hurricanes or storm-induced turbidity currents.

If exchange between the coastal ocean and the deep ocean plays a significant role in biogeochemical cycles, how can these processes be included in models? Interdisciplinary modeling of coastal systems has received more attention than that of the open ocean (see the recent review by Wroblewski and Hoffman, 1989). These models have tended to focus on the local effects of a particular physical phenomenon. By limiting the spatial and temporal extent of the models, it is possible to resolve very small scales and to explicitly calculate the effect of nonlinear physical-biological interactions.

As was the case for the mixed-layer models, it is difficult to see how to scale up the detailed coastal models to larger scales. One possibility may be to use the results of a high-resolution coastal margin model to derive flux boundary conditions for a larger scale model.

A second approach is to parameterize the effect of continental boundaries in some way. The primary productivity maps of Berger et al. (1987), for example, provide some reassuring evidence that the main distinction between the continental margins and the open ocean can be represented in a simple manner. Berger et al. constructed their map by assuming that primary production was a function of only three variables: phosphate concentration at 100 m, latitude, and distance from land. The resulting map looks very similar to a map based on direct productivity observations (Figure 6).

### **Particle Cycling in the Water Column**

Relatively little work has been done on explicitly modeling the dynamics of particulate organic matter as it sinks through the water column. More commonly, the rate of decomposition is expressed as an empirical relation derived from sediment trap results. These rules generally express the decrease in organic carbon flux as an exponential or power law function of depth (e.g., Suess, 1980; Berger et al., 1987; Martin et al., 1987). However, the depth exponent varies over a wide range in different formulations. Bishop (1989) has suggested that the large spread may be due to differences in zooplankton feeding in different environments.

As mentioned in the first section, calcium carbonate dissolves at a greater depth than that at which organic matter decomposes. As a result, changes in the species of the organisms contributing the bulk of the particle flux out of the euphotic zone can affect the vertical distribution of carbon in the ocean. In models it is frequently assumed that all calcium carbonate dissolution occurs on the sea floor, but the extent to which dissolution may occur in the water column is unknown.

Models of particle cycling have also been constructed based on thorium isotope distributions. Thorium is very effectively scavenged by particles, and the disequilibrium between  $^{234}\text{Th}$  and its parent  $^{238}\text{U}$  gives a measure of the particle flux (e.g., Coale and Bruland, 1985). The isotope-based particle models have not yet been coupled to carbon cycle models.

### **Benthic Processes**

A small fraction (1–2%) of the export production escapes being regenerated in the water column and reaches the sea floor. Most of this organic matter is oxidized at the surface of the sediments, providing an energy source for benthic organisms, as well as a source of dissolved carbon and a sink for oxygen. Berger et al. (1989) estimate that about 30% of the increase in apparent oxygen utilization

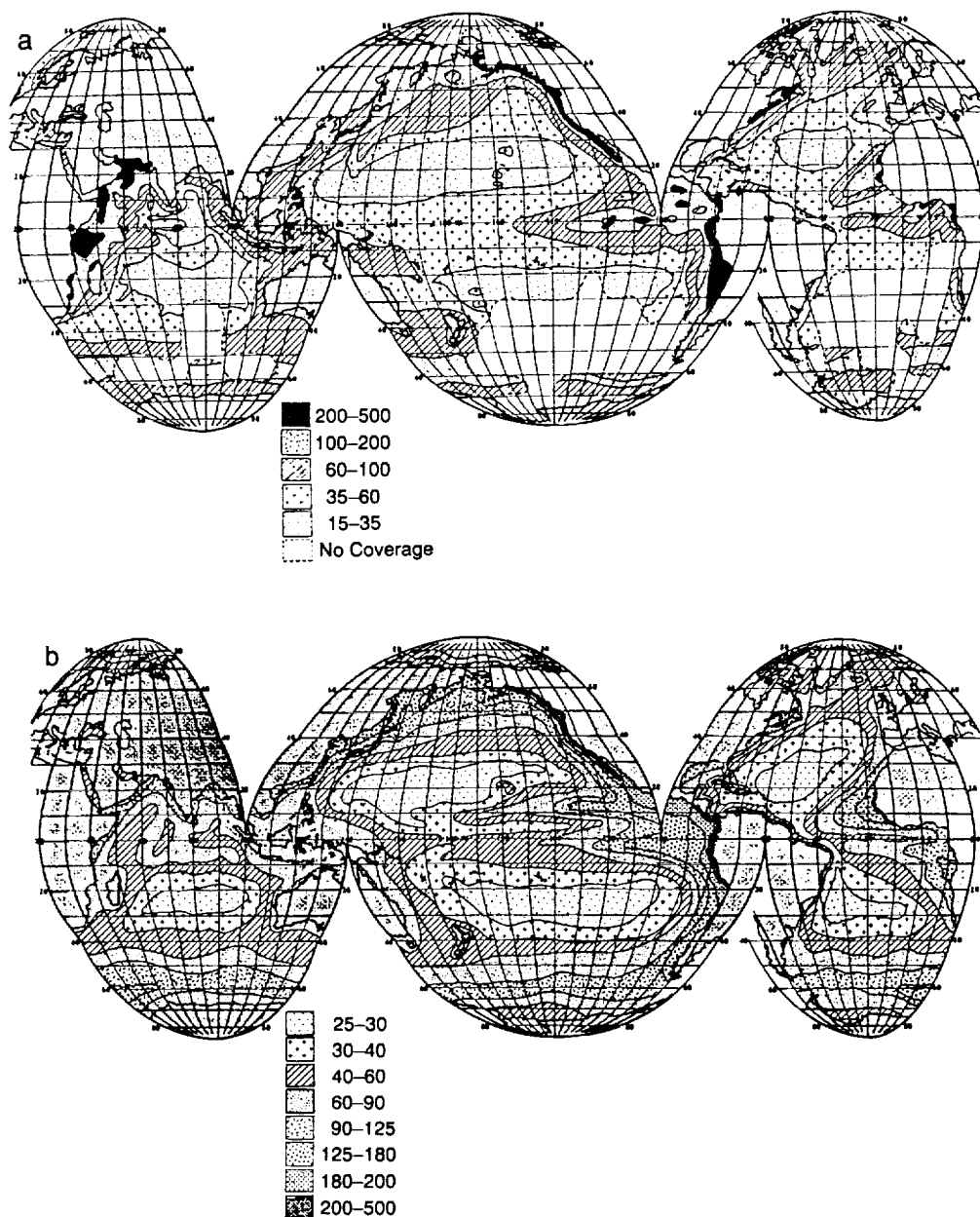


Figure 6. Annual primary production (PP) of the world's ocean, in  $\text{g C/m}^2/\text{yr}$ : (a) based on integrated productivity measurements compiled from the literature and (b) modeled using phosphate distributions, latitude, and distance from shore (adapted from Berger, 1989).

(AOU) with depth is the result of bottom respiration. The dynamics of the bottom boundary layer control the flux of regenerated carbon and nutrients from the sea floor back into the interior. Thus far this aspect of biogeochemical cycling in the ocean has received little attention in models.

### **Recommendations for Improved Coupled Physical-Biogeochemical Models**

The models discussed above have displayed a wide range of complexity and scale, from the two-box model in which the physics and the biology are each parameterized by a single term, to detailed mixed-layer models in which the trajectories of individual phytoplankton are resolved. The main difficulty in designing coupled physical-biogeochemical models that can be used to predict the system response to changing external forcing is determining the right compromises to make: We need a model that reproduces the essential interactions, and yet is simple enough to be feasible with finite computational resources. By "feasible" in this case I mean cheap enough to run that a variety of experiments and sensitivity tests can be performed.

The first problem we face is identifying these "essential interactions." The mixed-layer models have demonstrated the sensitivity of new production to physical-biological interactions occurring on small space and time scales (e.g., resonant interactions between variable winds and inertial currents, or adaptation of photosynthetic efficiency to variable light levels). These models suggest that the essential nature of the processes depends on the nonlinear interactions between physical and biological fields varying in space and time. On the other hand, simple empirical relations seem to capture the net effect of these small-scale effects with some success. The maps of Berger et al. (1987), for example, suggest that there is a simpler underlying structure to the coupled system we are trying to model.

Clearly, empirical relations are not enough if we are to construct a predictive model that can describe how the system changes in response to changing forcing. The trick to finding parameterizations that reflect the essence of the active mechanisms is to understand the system well enough to say which pieces are important and which are not. At present we can do this for few, if any, of the processes involved in oceanic biogeochemical cycles.

It is unusual for a modeler to step back from a calculation and ask: How much can I simplify my model and still reproduce the essential results? Such an exercise, however, could be a fruitful way to derive appropriate parameterizations. At present we too fre-

quently choose a particular simplification because it is "sensible" and feasible, rather than by demonstrating that the parameterization retains the important features of the complex system.

The terrestrial ecosystem models discussed by Running in this volume provide an example of the type of "devolutionary" model development I have in mind. The natural tendency in making models is to gradually increase their complexity in an effort to make the simulation more closely resemble the observations. In the terrestrial context, this tendency led to ecosystem models based on individual trees, whose physiology was modeled explicitly. As interest shifted to regional scales, it became apparent that this level of detail was unnecessary. In particular, Running has found that climate, leaf area index, and soil water capacity are sufficient to define the regional ecosystem. For the evergreen forests he considered, the time steps could be increased from one hour to one day for hydrologic processes, and to one year for carbon and nutrient cycling. In this case, the simplifications introduced could be justified by comparing the results to those of the more detailed model. Such a comparison, for example, showed that if the time step for hydrologic processes was increased beyond one day, the simulation was no longer adequate.

A similar exercise is necessary in the ocean if we are to have confidence in the results of coupled physical-biogeochemical models, which will of necessity be highly parameterized. A major stumbling block has been the lack of observations. The nonlinear nature of the interactions causes difficulties for the observationalist as well as the modeler; physical and biogeochemical variables are rarely measured at the same time and place, with the necessary resolution and duration. Focused mesoscale studies involving close collaboration between observers and modelers, physicists and biologists, in a variety of regions are needed to develop high-resolution models for the individual regions. These detailed models can then be used to calibrate the simpler models appropriate for coupling to global ocean and atmosphere models.

The lack of observations also causes problems for the modeler, who needs data to calibrate the model parameters and to validate the model. In this regard it is important that the model produce as output an analog of something that is measurable. Satellite observations of ocean color, for example, provide a global, synoptic data set of some quantity related to primary productivity, but this is related to typical model output such as net export production in a complicated and unknown way.

Given the variety of evidence, from both models and observational studies, of the importance of episodic events in the biogeochemical

system, more work is required on including a stochastic element in our models.

We need a different sort of physical model as well. Of primary importance is a more realistic treatment of the upper ocean. While an accurate prediction of the properties of the mixed layer is of obvious importance to the biology, it is also a necessity for improved modeling of the ventilation processes responsible for the physically mediated transport of carbon into the ocean interior. A better representation of the upper ocean is also necessary to allow direct coupling of atmospheric and oceanic GCMs without the introduction of ad hoc "flux corrections" (see Rintoul, this volume).

At the moment, ocean models used for climate studies lie at one of two extremes: simple box models, in which the dynamics are reduced to a small number of exchange coefficients; and GCMs, which are computationally intensive but still do not resolve the mixed layer very well. There is a need for a model of intermediate complexity, particularly for consideration of climate change on decadal time scales. Such a model should include a well-resolved mixed layer and thermocline, with coarser resolution in the deep sea. In addition, the mixing parameterization used should more accurately reflect the nature of the physical processes responsible for mixing in the ocean (e.g., internal mixing along isopycnals, vertical mixing dependent on vertical shear or stability). This type of model would more accurately represent ocean ventilation processes than present GCMs. It would also be more appropriate for coupling to biological models.

Ideally, we would like a model that is consistent with everything we know from past observations and physical reasoning, and that can be integrated forward in time to predict how things change. Thus, we would like a model that is capable of assimilating a wide variety of data: satellite measurements of ocean color, sea surface temperature and sea surface height, observations of *f*-ratio or community structure, an estimate of the meridional heat flux, a float trajectory or current meter measurement, etc. We would also like the model to determine the things we know least well from information on the things we do know well. For example, one might use interior observations of temperature and salinity to constrain estimates of the heat and fresh water flux at the sea surface (e.g., see Tziperman, this volume). Much progress is currently being made in the field of data assimilation or optimal estimation in meteorology and oceanography. These methods are in a sense a marriage of the prognostic and diagnostic methods, and may hold promise for the biogeochemical problem as well, although there is a long way to go in the development of such models.

## Acknowledgments

I would like to thank the organizers of the Global Change Institute for a stimulating and enjoyable conference. Part of this paper was prepared while I was a visiting scientist in the Atmospheric and Oceanic Sciences Program at Princeton University. I thank the AOS program for its hospitality and, in particular, J. Sarmiento, R. Toggweiler, and K. Bryan for many interesting discussions during my stay.

## References

- Bacastow, R., and E. Maier-Reimer. 1990. Ocean-circulation model of the carbon cycle. *Climate Dynamics* 4, 95-126.
- Berger, W.H. 1989. Global maps of ocean productivity. In *Productivity of the Ocean: Present and Past* (W.H. Berger, V.S. Smetacek, and G. Wefer, eds.), Dahlem Conference, John Wiley and Sons, Chichester, England, 429-456.
- Berger, W.H., V.S. Smetacek, and G. Wefer. 1989. Ocean productivity and paleoproductivity—an overview. In *Productivity of the Ocean: Present and Past* (W.H. Berger, V.S. Smetacek, and G. Wefer, eds.), Dahlem Conference, John Wiley and Sons, Chichester, England, 1-34.
- Berger, W.H., K. Fisher, C. Laai, and G. Wu. 1987. Ocean productivity and organic carbon flux. Part I. Overview and maps of primary productivity and export production. SIO Reference 87-30, University of California, San Diego.
- Bishop, J.K.B. 1989. Regional extremes in particulate matter composition and flux: Effects on the chemistry of the ocean interior. In *Productivity of the Ocean: Present and Past* (W.H. Berger, V.S. Smetacek, and G. Wefer, eds.), Dahlem Conference, John Wiley and Sons, Chichester, England, 117-138.
- Bolin, B., A. Björkström, and B. Moore. 1989. Uptake by the Atlantic Ocean of excess atmospheric carbon dioxide and radiocarbon. In *Understanding Climate Change* (A. Berger, R.E. Dickinson, and J.W. Kidson, eds.), Geophysical Monograph 52, American Geophysical Union, Washington, D.C., 57-78.
- Boyle, E.A. 1988. The role of vertical chemical fractionation in controlling late quaternary atmospheric carbon dioxide. *Journal of Geophysical Research* 93, 15701-15714.
- Brewer, P.G. 1986. What controls the variability of carbon dioxide in the surface ocean? A plea for complete information. In *Dynamic Processes in the Chemistry of the Upper Ocean* (J.D. Burton, P.G. Brewer, and R. Chesselet, eds.), Plenum Press, New York, 215-231.
- Broecker, W.S. 1982. Ocean chemistry during glacial time. *Geochimica et Cosmochimica Acta* 46, 1689-1705.



- Broecker, W.S., and T.-H. Peng. 1982. *Tracers in the Sea*. Eldigio Press, Lamont-Doherty Geophysical Observatory, Palisades, New York, 690 pp.
- Broecker, W.S., and T.-H. Peng. 1986. Global carbon cycle, 1985: Glacial to interglacial changes in the operation of the carbon cycle. *Radiocarbon* 28, 309-327.
- Broecker, W.S., and T. Takahashi. 1984. Is there a tie between atmospheric CO<sub>2</sub> content and ocean circulation? In *Climate Processes and Climate Sensitivity* (J.E. Hansen and T. Takahashi, eds.), Geophysical Monograph 29, American Geophysical Union, Washington, D.C., 314-326.
- Bryan, F. 1987. Parameter sensitivity of primitive equation ocean circulation models. *Journal of Physical Oceanography* 17, 970-985.
- Bryan, K. 1979. Models of the world ocean. *Dynamics of the Atmosphere and Oceans* 3, 327-338.
- Coale, K.H., and K.W. Bruland. 1985. Th-234/U-238 disequilibria within the California Coastal Current. *Limnology and Oceanography* 30(1), 22-33.
- Codispoti, L.A., G.E. Friederick, R.L. Iverson, and D.W. Hood. 1982. Temporal changes in the inorganic carbon system of the southeastern Bering Sea during spring 1980. *Nature* 296, 242-244.
- Craig, H. 1957. The natural distribution of radiocarbon and the exchange time of carbon dioxide between atmosphere and sea. *Tellus* 9, 1-17.
- Dymond, J., and M. Lyle. 1985. Flux comparisons between sediments and sediment traps in the eastern tropical Pacific: Implications for atmospheric CO<sub>2</sub> variations during the Pleistocene. *Limnology and Oceanography* 30, 699-712.
- Emerson, S. 1985. Organic carbon preservation in marine sediments. In *Carbon Dioxide and the Carbon Cycle: Natural Variations Archean to Present* (E.T. Sundquist and W.S. Broecker, eds.), Geophysical Monograph 32, American Geophysical Union, Washington, D.C., 78-88.
- Esalas, W.E., G.C. Feldman, C.R. McClain, and J.A. Elrod. 1986. Monthly satellite-derived phytoplankton pigment distribution for the North Atlantic Ocean basin. *Eos* 67, 835-837.
- Fasham, M.J.R. 1985. Flow analysis of materials in the marine euphotic zone. *Canadian Bulletin of Fisheries and Aquatic Science* 213, 139-162.
- Fasham, M.J.R., H.W. Ducklow, and S.M. McKelvie. 1990. A nitrogen-based model of plankton dynamics in the oceanic mixed layer. *Journal of Marine Research* 48(3), 591-639.

- Garçon, V., and J.F. Minster. 1988. Heat, carbon and water fluxes in a 12-box model of the world ocean. *Tellus* 40B, 161-177.
- Houghton, J.T., G.J. Jenkins, and J.J. Ephraums (eds.). 1990. *Climate Change: The IPCC Scientific Assessment*, Intergovernmental Panel on Climate Change and Cambridge University Press, Cambridge, 365 pp.
- Jenkins, W.J., and J.C. Goldman. 1985. Seasonal oxygen cycling and primary production in the Sargasso Sea. *Journal of Marine Research* 43, 465-491.
- Keeling, C.D. 1968. Carbon dioxide in surface ocean waters, 4: Global distribution. *Journal of Geophysical Research* 73, 4,543-4,553.
- Keir, R. 1988. On the late Pleistocene ocean geochemistry and circulation. *Paleoceanography* 3, 443-445.
- Kiefer, D.A., and J.N. Kremer. 1981. Origins of vertical patterns of phytoplankton and nutrients in the temperate, open ocean: A stratigraphic hypothesis. *Deep Sea Research* 28, 1087-1105.
- Klein, P., and B. Coste. 1984. Effects of wind-stress variability on nutrient transport into the mixed layer. *Deep Sea Research* 31, 21-37.
- Knox, F., and M. McElroy. 1984. Changes in atmospheric CO<sub>2</sub>: Influence of biota at high latitudes. *Journal of Geophysical Research* 89, 4629-4637.
- Lin, C.A. 1988. A mechanistic model of isopycnal diffusion in the ocean. *Climate Dynamics* 2, 165-172.
- Liss, P.S., and L. Merlivat. 1986. Air-sea gas exchange rates: Introduction and synthesis. In *The Role of Air-Sea Exchange in Geochemical Cycling* (P. Buat-Menard, ed.), D. Reidel, Hingham, Massachusetts, 113-127.
- Martin, J.H., G.A. Knauer, D.M. Karl, and W.H. Broenkow. 1987. VERTEX: Carbon cycling in the northeast Pacific. *Deep Sea Research* 34, 267-285.
- Martin, J.H., and S.E. Fitzwater. 1988. Iron deficiency limits phytoplankton growth in the north-east Pacific subarctic. *Nature* 331, 341-343.
- Martin, J.H., and R.M. Gordon. 1988. Northeast Pacific iron distribution in relation to phytoplankton productivity. *Deep Sea Research* 34, 267-285.
- Martin, J.H., R.M. Gordon, and S.E. Fitzwater. 1990. Iron in Antarctic waters. *Nature* 345, 156-158.
- Metzl, N., B. Moore, and A. Poisson. 1990. Resolving the intermediate and deep advective flows in the Indian Ocean by using temperature, salinity, oxygen and phosphate data: The interplay of biogeochemical and geophysical tracers. *Palaeogeography, Palaeoclimatology, Palaeoecology (Global and Planetary Change Section)* 89, 81-111.

- Michaels, A.F., and M.W. Silver. 1988. Primary production, sinking fluxes and the microbial foodweb. *Deep Sea Research* 35, 473-490.
- Musgrave, D.L., J. Chou, and W.J. Jenkins. 1988. Application of a model of upper-ocean physics for studying seasonal cycles of oxygen. *Journal of Geophysical Research* 93, 15679-15700.
- Neftel, A., H. Oeschger, J. Schwander, B. Stauffer, and R. Zimbrum. 1982. Ice core measurements give atmospheric CO<sub>2</sub> content during the past 40,000 years. *Nature* 295, 222-223.
- Nelson, D.M., J.J. McCarthy, T.M. Joyce, and H.W. Ducklow. 1989. Enhanced near-surface nutrient availability and new production resulting from the frictional decay of a Gulf Stream warm-core ring. *Deep Sea Research* 36, 705-714.
- Nydal, R. 1968. Further investigation on the transfer of radiocarbon in nature. *Journal of Geophysical Research* 75, 2271-2278.
- Oeschger, H., U. Siegenthaler, U. Schotterer, and A. Gugelmann. 1975. A box diffusion model to study the carbon dioxide exchange in nature. *Tellus* 27, 168-192.
- Price, J.R., R. A. Weller, and R. Pinkel. 1986. Diurnal cycling: Observations and models of the upper ocean response to diurnal heating, cooling, and wind mixing. *Journal of Geophysical Research* 91, 8411-8427.
- Redfield, A.C., B.H. Ketchum, and F.A. Richards. 1963. The influence of organisms on the composition of sea-water. In *The Sea, Volume 2: The Composition of Sea Water: Comparative and Descriptive Oceanography* (M.N. Hill, ed.), Interscience Publishers, New York, 26-77.
- Revelle, R., and H. E. Suess. 1957. Carbon dioxide exchange between atmosphere and ocean and the question of an increase of atmospheric CO<sub>2</sub> during the past decade. *Tellus* 9, 18-27.
- Riley, G.A. 1951. Oxygen, phosphate and nitrate in the Atlantic Ocean. *Bulletin of the Bingham Oceanographic Collection* 12(3), 169 pp.
- Rintoul, S.R., and C. Wunsch. 1991. Mass, heat, oxygen, and nutrient fluxes and budgets in the North Atlantic Ocean. *Deep Sea Research* 38(1A), S355-S377.
- Rowe, G.T., S. Smith, P. Falkowski, T. Whittedge, R. Theroux, W. Phoel, and H. Ducklow. 1986. Do continental shelves export organic matter? *Nature* 324, 559-561.
- Sarmiento, J.L., and K. Bryan. 1982. An ocean transport model for the North Atlantic. *Journal of Geophysical Research* 87, 394-408.
- Sarmiento, J.L., and J.R. Toggweiler. 1984. A new model for the role of the oceans in determining atmospheric pCO<sub>2</sub>. *Nature* 308, 621-624.

- Sarmiento, J.L., J.R. Toggweiler, and R. Najjar. 1988. Ocean carbon cycle dynamics and atmospheric  $p\text{CO}_2$ . *Philosophical Transactions of the Royal Society of London A* 325, 3–21.
- Sarmiento, J.L., M.J.R. Fasham, U. Siegenthaler, R. Najjar, and R. Toggweiler. 1989. *Models of Chemical Cycling in the Oceans: Progress Report II*. Ocean Tracers Laboratory Technical Report No. 6, Princeton University, Princeton, New Jersey.
- Sarmiento, J.L., J.C. Orr, and U. Siegenthaler. 1990. A perturbation simulation of  $\text{CO}_2$  uptake in an ocean general circulation model. *Journal of Geophysical Research*.
- Schlitzer, R. 1988. Modeling the nutrient and carbon cycles of the North Atlantic. Part I: Circulation, mixing coefficients and heat fluxes. *Journal of Geophysical Research* 93, 699–723.
- Shackleton, N.J., and N.G. Pisias. 1985. Atmospheric carbon dioxide, orbital forcing and climate. In *Carbon Dioxide and the Carbon Cycle, Archean to Present* (E.T. Sundquist and W.S. Broecker, eds.), American Geophysical Union, Washington, D.C., 313–318.
- Shulenberg, E., and J.L. Reid. The Pacific shallow oxygen maximum, deep chlorophyll maximum and primary productivity, reconsidered. *Deep Sea Research* 28, 901–920.
- Siegenthaler, U., and T. Wenk. 1984. Rapid atmospheric  $\text{CO}_2$  variations and ocean circulation. *Nature* 308, 624–625.
- Suess, E. 1980. Particulate organic carbon flux in the oceans—surface productivity and oxygen utilization. *Nature* 288, 260–263.
- Sugimura, Y., and Y. Suzuki. 1988. A high-temperature catalytic oxidation method for the determination of non-volatile dissolved organic carbon in seawater by direct injection of liquid samples. *Marine Chemistry* 24, 105–131.
- Sundquist, E.T. 1985. Geological perspectives on carbon dioxide and the carbon cycle. In *Productivity of the Ocean: Present and Past* (W.H. Berger, V.S. Smetacek, and G. Wefer, eds.), Dahlem Conference, John Wiley and Sons, Chichester, England, 99–110.
- Tans, P.P., I.Y. Fung, and T. Takahashi. 1990. Observational constraints on the global atmospheric carbon budget. *Science* 247, 1431–1438.
- Toggweiler, J.R. 1989. Is the downward dissolved organic matter (DOM) flux important in carbon transport? In *Productivity of the Ocean: Present and Past* (W.H. Berger, V.S. Smetacek, and G. Wefer, eds.), Dahlem Conference, John Wiley and Sons, Chichester, England, 65–85.
- Vezina, A.F., and T. Platt. 1987. Small-scale variability of new production and particulate fluxes in the ocean. *Canadian Journal of Fisheries and Aquatic Science* 44, 198–205.

- 
- Volk, T. 1989. The changing patterns of  $p\text{CO}_2$  between ocean and atmosphere. *Global Biogeochemical Cycles* 3, 179–189.
- Volk, T., and M.J. Hoffert. 1985. Ocean carbon pumps: Analysis of relative strengths and efficiencies in ocean-driven atmospheric  $\text{CO}_2$  changes. In *Carbon Dioxide and the Carbon Cycle, Archean to Present* (E.T. Sundquist and W.S. Broecker, eds.), American Geophysical Union, Washington, D.C., 313–318.
- Walsh, J.J., G.T. Rowe, R.L. Iverson, and C.P. McRoy. 1981. Biological export of shelf carbon is a sink of the global  $\text{CO}_2$  cycle. *Nature* 291, 196–201.
- Williams, P.J.leB., and B. von Bodungen. 1989. Export productivity from the photic zone. In *Productivity of the Ocean: Present and Past* (W.H. Berger, V.S. Smetacek, and G. Wefer, eds.), Dahlem Conference, John Wiley and Sons, Chichester, England, 99–115.
- Woods, J.D., and W. Barkmann. 1986. A Lagrangian mixed layer model of Atlantic  $18^\circ\text{C}$  water formation. *Nature* 391, 574–576.
- Woods, J.D., and R. Onken. 1982. Diurnal variation and primary production in the ocean—preliminary results of a Lagrangian ensemble model. *Journal of Plankton Research* 4, 735–756.
- Wroblewski, J. S., and E.E. Hoffman. 1989. U.S. interdisciplinary modeling studies of coastal-offshore exchange processes: Past and future. *Progress in Oceanography* 23, 65–100.
- Wroblewski, J.S., J.L. Sarmiento, and G.R. Flierl. 1988. An ocean basin scale model of plankton dynamics in the North Atlantic. 1. Solutions for the climatological oceanographic conditions in May. *Global Biogeochemical Cycles* 2(3), 199–218.

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100

100